# MONSOON-OCEAN INTERACTIONS

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The Asian-Australian monsoon (AAM) covers the entire Indo-Pacific warm pool region (e.g., Ramage 1971). Thus, the maritime monsoon is an integral and essential component of the AAM system. Large-scale variations of the monsoon circulation and the underlying oceanic mixed layer have been observed on time scales ranging from intraseasonal to interdecadal. The nature of monsoon-ocean interactions may vary with time scale. The coupling on the intraseasonal and interdecadal time scales is discussed elsewhere by other presenters in this workshop. Here we shall focus on monsoon-ocean interactions on interannual time scales. The covariability of the atmosphere-ocean system in the AAM region on this time scale is strongly modulated by the evolution of El Niño-Southern Oscillation (ENSO) events in the central and eastern tropical Pacific. Of particular interest is the response of the monsoon flows to local and remote sea surface temperature (SST) changes that emerge during ENSO, as well as the impact of monsoon fluctuations on ENSO development.

Detailed accounts of the myriad atmospheric and oceanic features associated with ENSO, as well as the mechanisms contributing to ENSO variability, have been given by Rasmusson (1985), Enfield (1989), Philander (1990), Glantz *et al.* (1991), Cane (1992), Wallace *et al.* (1998) and Neelin *et al.* (1998), among others. The principal goal of the present review is to offer a synopsis of the monsoonal features in the South Asian and East Asian-Western Pacific sectors in different stages of the ENSO cycle, physical mechanisms that contribute to covariability between ENSO and the monsoon system, the effects of monsoon anomalies on the oceanic temperature and circulation fields, the local oceanic feedbacks to monsoon variability, and the implications of such feedbacks on the subsequent development of ENSO.

Much of the material in this report is excerpted from a more comprehensive review of monsoon-ENSO interactions by Lau and Wang (2005), to which the interested readers are referred.

## 1. Precipitation Anomalies in the AAM Region during ENSO Events

The typical evolution of precipitation anomalies in the AAM region in various phases of the ENSO cycle has been documented by Ropelewski and Halpert (1987) using station records. Various other empirical studies on the impacts of ENSO on the monsoon rainfall intensity over the Indian subcontinent, East Asia and Australia have also been reviewed recently by Webster *et al.* (1998) and Wang *et al.* (2003).

To gain an overview of the development of the anomalous rainfall pattern through different stages of ENSO, the composite of the seasonally averaged precipitation field over the cold La Niña events of 1988 and 1998 have been subtracted from the corresponding composite over the warm El Niño events of 1982, 1991 and 1997. The patterns thus obtained (hereafter referred to as 'warm-minus-cold composites') are displayed in Fig. 1, for the period extending from the boreal summer of the year when the events initiated ('Year 0') to the summer of the following year ('Year 1'). We shall henceforth refer to a specific time period within the ENSO time frame by grouping the first letter of the months in that period, followed by the year(s) in parentheses. For instance, the five panels in Fig. 1 correspond to the JJA(0), SON(0), DJF(0/1), MAM(1) and JJA(1) seasons. These charts have been constructed using the dataset produced by the Global Precipitation Climatology Project (GPCP; see Huffman *et al.* 1997), which incorporates measurements by both rain gauges and satellites in the era from 1979 to present.

Through much of the JJA(0)-DJF(0/1) period (Figs. 1a-1c), the precipitation patterns in the equatorial zone are dominated by negative anomalies over the Indonesian Archipelago and eastern Indian Ocean, and by positive anomalies from ~150$^{\circ}$E to the dateline. These features are indicative of the eastward displacement of the Walker Circulation during warm ENSO events. Over the Arabian Sea/India/Bay of Bengal region, below-normal rainfall prevails during the summer and autumn of Year(0) (Figs. 1a-1b). This deficiency of Indian monsoon rainfall during the summer of warm events is a well-known phenomenon (e.g., Rasmusson and Carpenter 1983; Shukla and Paolino 1983). Comparison between Figs. 1a and 1e suggests that the summertime precipitation anomalies over the Arabian Sea and Bay of Bengal tend to change sign from Year(0) to Year(1). During SON(0) and DJF(0/1) (Figs. 1b-1c), the precipitation changes over the equatorial and southern Indian Ocean are characterized by dry conditions in the east, and wetness in the west. This rainfall pattern is evidently related to a recurrent mode of SST variability in the Indian Ocean basin, which also exhibits distinct east-west contrasts during the northern autumn season.

The most prominent precipitation anomalies in the East Asian and Australian monsoon regions appear in DJF(0/1) (Fig. 1c). Below-normal rainfall is observed over the Philippines and the nearby oceans, as well as northern Australia. An elongated wet zone is also seen to extend northeastward from southern China to the waters south of Japan. The dry anomaly in the vicinity of the Philippines is first established in SON(0) over the South China Sea (Fig. 1b). This feature migrates eastward with time, with its main center being located over the tropical western Pacific in MAM(1) (Fig. 1d). Remnants of the wet anomaly over southern China and Japan are still discernible in MAM(1).

## 2. ENSO-related Variability in the Indian Ocean Basin

### *a. Atmospheric and SST Anomalies*

The typical atmospheric and oceanic changes in the Indian Ocean (IO) sector during ENSO episodes are illustrated in Fig. 2, which shows the warm-minus-cold composites of the observed 850 hPa vector wind and SST fields for the JJA(0), SON(0) and DJF(0/1) seasons. These patterns are based on composites over the six warm events of 1957, 1965, 1972, 1982, 1991 and 1997, and the six cold events of 1955, 1970, 1973, 1975, 1988 and 1998, and have been constructed using the reanalyses produced by the National Centers for Environmental Prediction (NCEP).

The most coherent atmospheric signal in the pattern for JJA(0) (Figs. 2a) is the anticyclonic 850 hPa circulation anomaly that prevails over the Arabian Sea and the surrounding land areas. This feature is seen to extend towards the Bay of Bengal and Indochina during the SON(0) season (Figs. 2c). The easterly wind anomalies over much of the northern IO that accompany the anticyclone oppose the climatological westerlies over this region, and is indicative of below-normal intensity of the summer monsoon circulation over South Asia during warm ENSO events. Also evident in the composite patterns for the northern summer and fall seasons (Figs. 2b and 2d) is the emergence of warm SST anomalies in both the Arabian Sea and Bay of Bengal. As noted in Lau and Nath (2000, 2003), two factors contribute to these SST changes. First, reduction in the monsoon intensity during warm ENSO episodes is accompanied by lowered wind speeds over these oceanic regions, which result in less latent and sensible heat loss to the atmosphere. Secondly, the decreased amount of cloud cover due to the generally dryer conditions in these areas (see Figs. 1a-1b) leads to more heating of the ocean surface by incoming solar radiation. The observed SST increase near the Somali and Arabian coasts could also be partially caused by the reduced oceanic upwelling associated with weakened monsoon flows.

Another noteworthy SST signal in the JJA(0) and SON(0) seasons is the cold anomaly that develops off the Sumatra-Java coasts. This feature is collocated with low-level southeasterly or easterly wind anomalies, which are parallel to the local climatological circulation (e.g., see Lau and Nath 2000). Budget analysis performed by Lau and Nath (2003) indicates that the increased surface wind speeds in this region lead to increased latent and sensible heat loss from the ocean, as well as deepening of the local oceanic mixed layer. Both effects are conducive to SST cooling. The stronger upwelling driven by the intensified winds along the shores of Sumatra and Java, and by anomalous easterlies in the eastern equatorial IO, could further enhance the observed cold SST anomaly in those sites. This oceanic cooling in the eastern IO is in distinct contrast with the warming in the western portion of the basin as described in the preceding paragraph. The occurrence of this zonal asymmetric SST anomaly pattern, which is most evident in SON(0), has been noted by Webster *et al.* (1999) and Saji *et al.* (1999). A corresponding east-west contrast in the precipitation field is also discernible in the same season (see Fig. 1b). These investigators have attributed this mode of variability mostly to processes operating within the IO sector. However, the appearance of the same pattern in the ENSO composites shown in Figs. 2d indicates that the remote forcing from the tropical Pacific could also influence the SST field in the IO.

During the DJF(0/1) period (Figs. 2f), the cold SST anomaly in the tropical eastern IO is no longer discernible. The climatological low-level flow over this region switches from easterly to westerly in this season (e.g., Schott and McCreary 2001), so that the easterly wind anomalies (Fig. 2e) would lead to reduction of both wind speed and heat loss from the ocean. The below-normal rainfall in this area (Fig. 1c) results in decreased cloud amounts and increased incoming solar radiation. Both effects contribute to warming in the eastern IO. The SST anomalies in other parts of the IO remain to be positive in this season, with notable amplification in the Bay and Bengal and the central IO between 20$^{o}$S and 30$^{o}$S. These two sites are overlain by wind anomalies that oppose the local climatological circulation, which is oriented southwestward over the Bay of Bengal, and northwestward over southern IO during the DJF season. The resulting decrease in wind speed and oceanic heat loss are hence consistent with the more pronounced SST warming over these areas. The basin-wide atmospheric circulation anomaly in DJF(0/1) (Fig. 2e) is characterized by strong easterlies along the tropical IO, and a pair of anticyclonic cells straddling the Equator, with centers located over the northwestern Australia and the South China Sea. Composite SST patterns for the MAM(1) and JJA(1) seasons, as shown in Alexander *et al.* (2004), indicate that the principal warm SST anomalies in the IO basin persist through the northern summer season of Year(1).

### *b. Atmospheric Response to Anomalous Tropical Heating*

We next evaluate the extent to which the atmospheric wind anomalies depicted in Fig. 2 may be attributed to remote forcing by ENSO-related precipitation changes in the tropical zone. By invoking analytic solutions presented by Matsuno (1966) and Gill (1980) for tropical circulations induced by heating, several investigators (e.g., Chen and Yen 1994; Kawamura 1998; Lau and Nath 2000; Wang *et al.* 2003) have interpreted the low-level anticyclones over South Asia and southern IO (see left panels of Fig. 2) as Rossby-wave responses to anomalous cooling over the Indonesian Archipelago and the equatorial western Pacific. The latter heat sink is in turn linked to the eastward displacement of the Walker Circulation during warm ENSO events, which results in below-normal precipitation in the equatorial zone between 90°E and 150°E (Figs. 1a-1c). The effects of the altered diabatic forcing in this region on the atmospheric flow pattern have been demonstrated by Wang *et al.* (2003) and Lau *et al.* (2004) using solutions of stationary wave models for the JJA(0) and DJF(0/1) seasons, respectively.

### *c. Atmosphere-Ocean Feedbacks in the IO Basin*

The cumulative evidence presented in Figs. 1-2 and the stationary wave model solutions discussed in Section 2b highlight the following chain of processes linking ENSO events in the tropical Pacific to SST variations in the IO basin: eastward displacement of the Walker Circulation during warm ENSO episodes, reduced precipitation and latent heat release over Indonesia and tropical western Pacific, generation of atmospheric Rossby-wave responses to the northwest and southwest of the heat sink, and atmospheric driving of the SST field in the IO sector through modulation of surface latent and radiative fluxes as well as ocean currents. Hence the atmospheric circulation serves as a 'bridge' communicating the ENSO forcing in DTEP to oceanic changes elsewhere (Klein *et al.* 1999).

Further diagnoses by Lau and Nath (2000, 2003) and Lau *et al.* (2004) using general circulation model (GCM) experiments demonstrate that the SST anomalies induced by the atmospheric bridge mechanism could in turn feed back on the atmospheric circulation. In particular, these model results imply that such feedbacks lead to an increase in the summer monsoon rainfall over South Asia in Year(1) of warm ENSO events. This perturbation is in opposition to that observed in Year(0), when the precipitation over the same region is below normal (compare Fig. 1a with Fig. 1e). This tendency for some monsoonal variations to switch polarity from one year to the next may be viewed as one facet of the Tropical Biennial Oscillation (e.g., see Meehl 1997). The model evidence described here indicate that biennial changes of the South Asian monsoon may partially be the consequence of the following chain of processes: remote responses to ENSO forcing in JJA(0), generation of SST anomalies in the IO basin during SON(1)-DJF(0/1) (Figs. 2d and 2f) by the atmospheric bridge, and feedback of these oceanic perturbations on the atmosphere in MAM(1)-JJA(1).

## 3. ENSO-related Variability over East Asia, Australia and Western Pacific

### *a. Atmospheric and SST Anomalies*

The essential atmospheric and oceanic changes in the eastern portion of the AAM system during ENSO events are summarized in Fig. 3, which shows the warm-minus-cold composites of observed surface wind vector, SST, sea level pressure (SLP) and precipitation, for DJF(0/1) (left panels) and MAM(1) (right panels). Results are based on the same sets of six warm and six cold ENSO events used in constructing Fig. 2. The most prominent features over the western Pacific are organized about the pair of positive SLP anomalies over the Philippine Sea and off the eastern Australia seaboard (Figs. 3c and 3g). These pressure perturbations are coincident with anomalous anticyclonic flows at the surface (Figs. 3a and 3e) and below-normal rainfall (Figs. 3d and 3h). The stationary wave solutions discussed in Section 2b indicate that the two high pressure centers are responses to the anomalous heat sink over the tropical western Pacific.

As has been pointed out by Wang *et al.* (2000), the evolution of the anomalous SST pattern (Figs. 3b and 3f) is closely related to changes in the local surface circulation. For instance, the southwesterly wind anomalies to the west of the Philippine Sea anticyclone (hereafter abbreviated as PSAC) oppose the climatological northeasterly winter monsoon over that region. The reduction in the wind speed leads to suppression of oceanic heat loss and warm SST anomalies. Conversely, the intensification of the northeasterly monsoon flow by the wind anomalies to the east of this anticyclone brings about oceanic cooling in the subtropical northwestern Pacific. Besides its impact on the underlying SST field, the weakening of the dry winter monsoon over East Asia in DJF(0/1) is also accompanied by above-normal precipitation over South China and East China Sea (Fig. 3d). This wet anomaly is seen to persist through the following spring season (Fig. 3h).

### *b. Stages of PSAC Development*

#### *1) Atmospheric Preconditions in Early Autumn of Year(0)*

The composite analysis of Wang and Zhang (2002) indicates that the circulation in this phase of the ENSO cycle is characterized by an anomalous 850 hPa cyclone and 200 hPa anticyclone over the western North Pacific, as well as a cyclonic anomaly at 200 hPa over northeastern Asia. Intensified westerlies prevail within the $30^{o}$-$40^{o}$N zone over East Asia. These upper tropospheric signals are indicative of a deepened trough and southward displacement of the climatological jetstream over that region. A negative 500 hPa anomaly extends eastward from northern China to the western Pacific. The most prominent feature in the surface air temperature field is a cold anomaly that extends from the Asian interior to the North Pacific between $35^{o}$ and $50^{o}$N.

#### *2) Synoptic Development during PSAC Onset*

Wang and Zhang (2002) noted that, during the strong warm ENSO years, the SLP field over the South China Sea and Philippine Sea typically made a distinct transition to a persistent positive anomaly in the October-November period of Year(0). The most pronounced features just prior to such transitions are the anticyclonic wind pattern and dryness associated with a high pressure anomaly over the interior of the Asian land mass, cyclonic flow and wet conditions accompanying a low center over the Philippine Sea, and prevalent northerly wind anomalies over southeastern China, East China Sea and southeastern Japan. These atmospheric signals bear a strong similarity to the characteristic behavior of cold air outbreaks over this region. The continental high pressure anomaly migrates southeastward to the Philippine Sea, and leads to the establishment of anticyclonic flows and dry conditions in the latter area. Wang and Zhang (2002) also pointed out that the deepening of the upper level trough and below-normal air temperature over East Asia in the preceding months constitute a favorable environment for the incidence of cold air outbreaks and the subsequent PSAC formation. These investigators have further considered the effects of intraseasonal oscillations on the rather abrupt reversal of the wind, pressure and rainfall anomalies over the Philippine Sea in this onset stage, and the role of atmosphere-ocean interactions in the seasonal dependence of the amplitude of such oscillations.

#### *3) Air-Sea Feedbacks in MAM(1)*

The atmospheric impact of the SST changes associated with PSAC development has been evaluated by Lau *et al.* (2004) using GCM experiments. These model analyses indicate that the air-sea feedbacks attendant to the warm SST anomaly in the $10^{o}$-$20^{o}$N $110^{o}$-$140^{o}$E region (see Fig. 3f) leads to stronger cyclonic development and more abundant precipitation along the climatological rainbelt extending northeastward from the southern coast of China to the western Pacific during MAM(1). Conversely, the cold SST anomaly in $10^{o}$-$20^{o}$N $150^{o}$-$180^{o}$E is overlain by increased SLP and reduced rainfall.

## 4. Impact of AAM on ENSO

How ENSO affects the AAM is better understood than the influence of the AAM on ENSO. The Asian monsoon covers one-third of the area of the tropics, and is an interactive component of the climate system that can impact the slowly varying lower boundary conditions (Webster *et al.* 1998). In the following subsections, we shall separately consider the effects of the Indian and western North Pacific monsoons on ENSO.

### *a. Effects of Indian Monsoon*

Barnett (1984) noted that ENSO-related westerly anomalies in the western/central Pacific originate from the Indian Ocean. It has been speculated that the eastward propagation of the westerly anomalies from the Indian monsoon region to the Pacific Ocean could serve as a trigger for ENSO events. Moreover, Webster and Yang (1992) showed that when the broad-scale South Asian summer monsoon is stronger (weaker) than normal, the tropical Pacific trade winds are also stronger (weaker) than average. This result suggests that anomalous Indian monsoon could affect ENSO through changing the trade winds over the Pacific.

Model experiments of varying degrees of complexity indicate that fluctuations of the Indian monsoon are linked to changes in the trade winds over the equatorial Pacific, even in the absence of ENSO (Kirtman and Shukla 2000). By incorporating the effects of monsoon heating in the coupled model of Zebiak and Cane (1987), Chung and Nigam (1999) noted that the presence of monsoonal interaction results in a broader frequency distribution of ENSO variability and a population shift in amplitude towards stronger El Niño events. The model results presented by Kirtman and Shukla (2000) also suggest that a variable monsoon enhances ENSO variability, particularly three to six months after the summer monsoon ends. Further analyses of output from coupled GCM integrations by Wu and Kirtman (2004) indicate that ENSO-related monsoon variability has significant impacts on warm events but not the cold events. A weak (strong) monsoon enhances (weakens) an ongoing warm event.

### *b. Effects of the Western North Pacific Monsoon*

McBride and Nicholls (1983) showed that SST anomalies in the Indonesian region lead those in the eastern Pacific by 4-6 months. They speculated that air-sea interaction in the Indonesian region might in part be responsible for the turnabout of ENSO. We have noted in Section 3 that the atmosphere-ocean coupling in that sector is closely related to the anomalous anticyclone over the Philippine Sea (PSAC) during warm ENSO events. Wang and Zhang (2002) pointed out that a sharp increase in SLP over the Philippine Sea typically precedes the peak warming in the central equatorial Pacific by about one to three months. In conjunction with this pressure rise, strong anticyclonic surface wind anomalies appear over the Philippine Sea, with enhanced easterlies prevailing north of New Guinea (Figs. 3a and 3e). The sudden emergence of the equatorial easterly anomalies over the western equatorial Pacific may generate oceanic upwelling Kelvin waves that propagate along the equator into the eastern Pacific. The resulting vertical displacements of the thermocline would cause negative SST tendencies in the latter region, thereby leading to the turnabout from El Niño to La Niña conditions.

The positive feedback between the atmospheric Rossby wave and ocean mixed layer thermodynamics, which plays a key role in development of the PSAC, depends on the presence of climatological northeasterly trades (Wang *et al.* 2000). In the western subtropical North Pacific this favorable basic state exists only from late fall through the following early summer. This seasonal dependence implies that persistent western Pacific easterly anomalies occur preferentially in autumn and winter, thus favoring ENSO turnabout in the mid-winter. This temporal evolution also explains why strong El Niño episodes tend to decay quickly after their mature phase.

Kim and Lau (2000) have shown that a strong biennial tendency in the ENSO cycle could result from the occurrence of wind anomalies in the western Pacific six months after the SST anomaly peaks in the eastern Pacific. The analysis of Lau and Wu (2001) suggests that a stronger monsoon-ENSO relationship tends to occur in boreal summer immediately after a peak El Niño and before a pronounced La Niña (i.e., 1998, 1988, and 1983). Their result appears to support the idea that the summer monsoon could influence ENSO variability via the development of the PSAC. The wind forcing associated with the PSAC may be instrumental in enhancing the biennial component of the natural ENSO cycle.

It is worthwhile to mention that the above process operates mostly during strong warm events. There are three moderate warm events (1986-87, 1968-69, and 1976-77) with no reversal of the warming trend at the end of the year of El Niño development. The easterly anomalies in the equatorial western Pacific were not well established and did not persist in the boreal winter of these El Niño years. An important feature common to all three prolonged events is the insufficient strength of the central Pacific warming (central equatorial Pacific SST anomaly less than 1.5 standard deviation by the end of the El Niño year), whereas during the six strong events considered in Fig. 3, the corresponding anomalies were all above 1.5 standard deviations. This suggests that the strong warming in the equatorial central Pacific is probably necessary for the robust establishment of the western Pacific wind anomalies.

## 5. Summary and Discussion

The anomalous Walker Circulation induced by eastern Pacific warming has a descending branch over the vicinity of the maritime continent. The suppressed deep convection over that region (Fig. 1) can generate westward propagating, descending Rossby waves that are conducive to a weak Indian summer monsoon. However, the weakening monsoon can induce local warming in the Bay of Bengal (Fig. 2). As a result of this SST change, the Indian summer monsoon would tend to intensify. Thus, through local atmosphere-ocean interaction, the ENSO-induced monsoon anomalies in turn offset the 'direct' ENSO impacts. Another manifestation of monsoon-ocean interaction related to ENSO is the Indian Ocean dipole. During an eastern Pacific warming, the low-level anticyclonic pattern over the southern Indian Ocean enhances the cross-equatorial flow along the west coast of Sumatra (Fig. 2), which subsequently enhances coastal and equatorial upwelling and cools the SST off Sumatra. The resultant SST cooling in the equatorial eastern Indian Ocean, along with concomitant warming in the western Indian Ocean, forms a dipolar SST anomaly pattern (Saji *et al.* 1999).

Over the western North Pacific, the local air-sea interaction can maintain the ENSO-induced anticyclonic anomalies from mature phase of El Niño to its decay phase, thus leading to a prolonged impact of ENSO on East Asian summer monsoon (see Section 3). On interannual time scales, Wang *et al.* (2003) have proposed that local monsoon-ocean interaction is one of the fundamental driving mechanisms for the biennial variability of the AAM. Questions remain regarding how the anomalies over the western North Pacific and southern Indian Ocean interact to yield the biennial tendency of the continental scale monsoon system.

The surface conditions over ocean and land play a crucial role in determining the predictability of the climate system (Charney and Shukla 1981; Shukla 1998). The current state-of-the-art atmospheric GCMs, when forced by the observed sea surface temperature (SST) anomalies occurring in the strong 1996-1998 ENSO event, are not capable of generating the changes in Asian-Pacific summer monsoon rainfall observed in that period (Wang *et al.* 2004a,b). The models tend to yield positive correlations between the summer monsoon precipitation and the local SST anomaly, which are at odds with observations. These authors demonstrate that an atmospheric GCM, when coupled with an ocean model, simulates realistic SST-rainfall relationships. However, this atmospheric GCM fails to reproduce such relationships when it is forced by the same SST anomalies that are generated in the coupled run. This departure of the forced solution from the coupled solution is mainly due to the absence of atmospheric feedback in the former case. These model results suggest that ocean-atmosphere feedback processes are very important in the monsoon precipitation zones, and call for the reshaping of current strategies for predicting variations in monsoon climate. In particular, the traditional 'Tier-2' approach for forecasting these variations, which entails the forcing of an atmospheric GCM with prescribed SST conditions that have been predicted in a previous step, appears to be inadequate.

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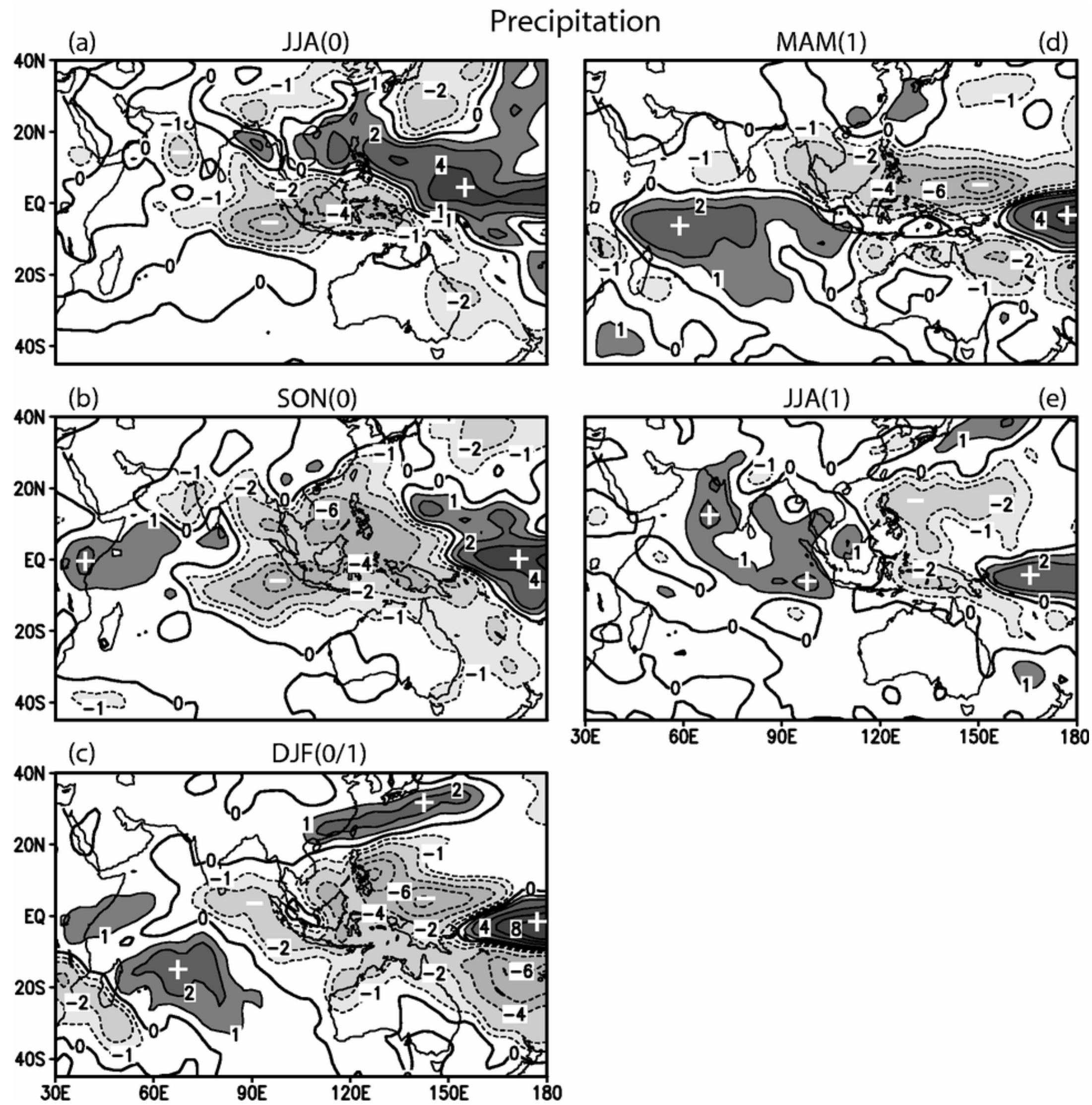


Fig. 1. Distributions of the warm-minus-cold composites of precipitation during (a) JJA(0), (b) SON(0), (c) DJF(0/1), (d) MAM(1) and (e) JJA(1), as computed using GPCP data for the warm ENSO events of 1982, 1991 and 1997 and the cold events of 1988 and 1998. Contour interval: 1 mm $d^{-1}$, with additional contours for -0.5 and +0.5 mm $d^{-1}$ being inserted. Solid and dashed contours indicate positive and negative values, respectively.

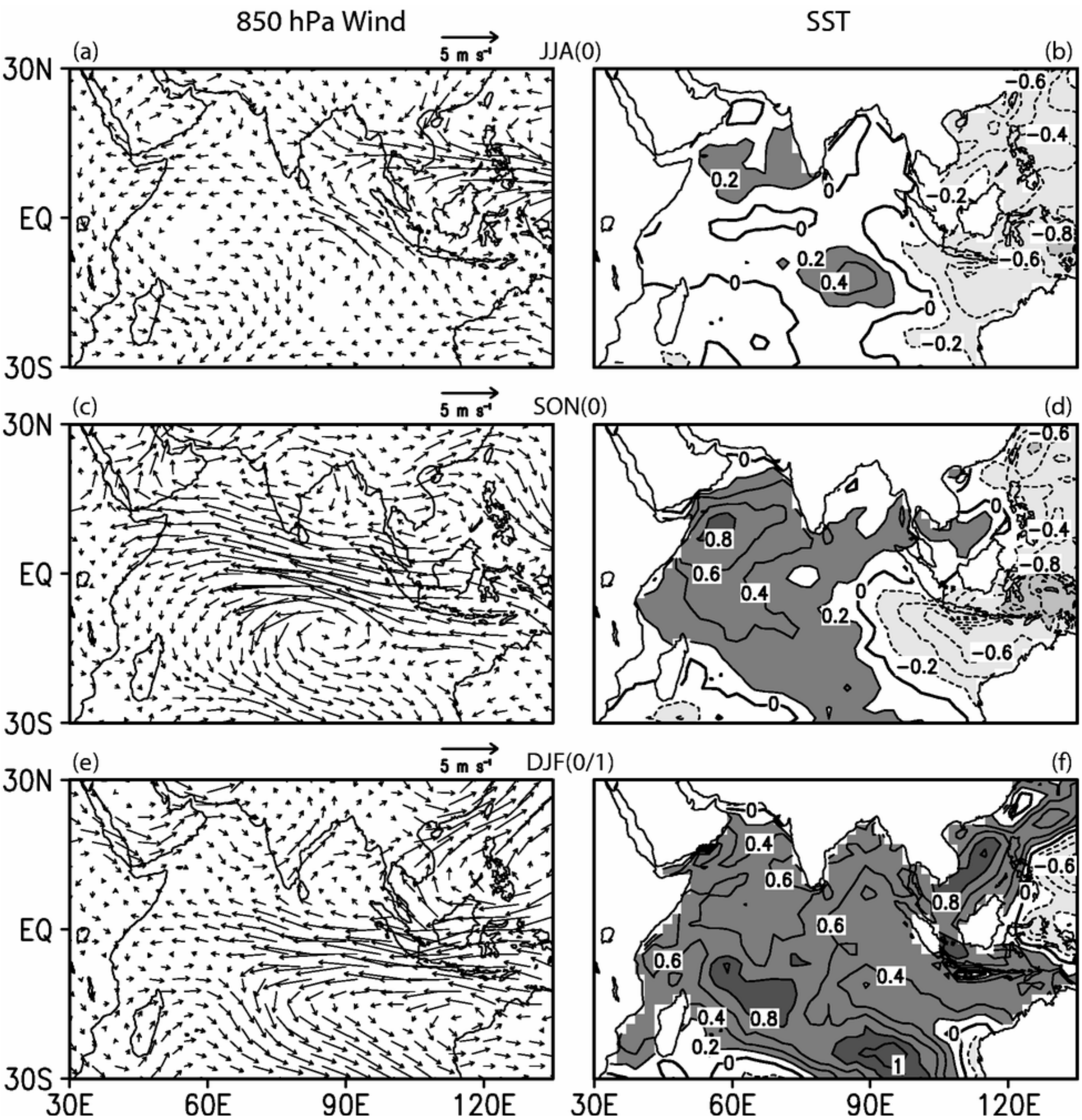


Fig. 2. Distributions of the warm-minus-cold composites of 850 hPa vector wind (left panels) and SST (right panels; contour interval: 0.2$^{o}$C) fields, for (a, b) JJA(0), (c, d) SON(0) and (e, f) DJF(0/1). Results are based on NCEP reanalysis data for six selected warm ENSO events and six cold events. Solid and dashed contours indicate positive and negative values, respectively.

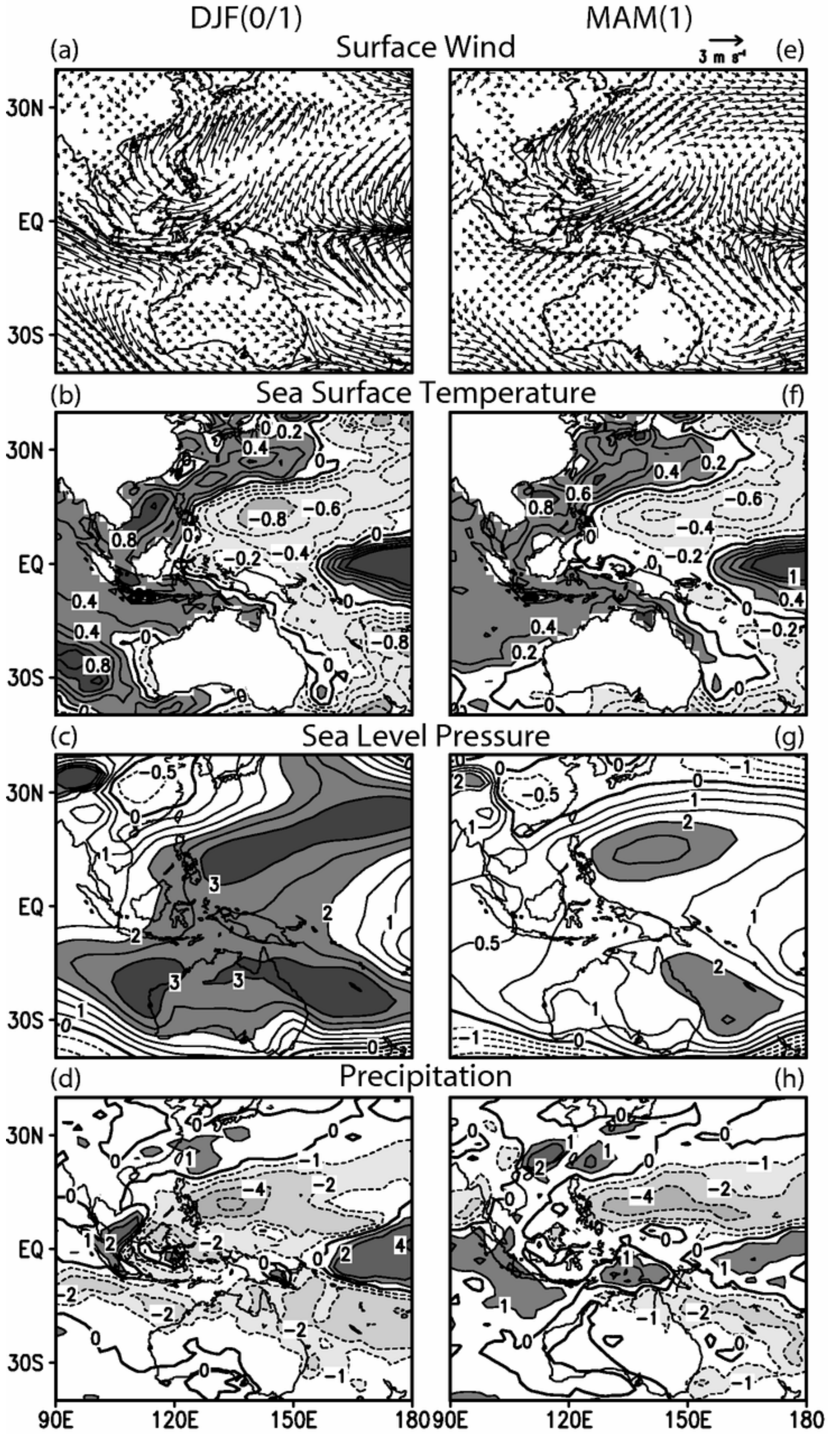


Fig. 3. Distributions of the warm-minus cold composites of (a, e) surface wind vector, (b, f) SST (contour interval: 0.2$^{o}$C), (c, g) SLP (contour interval: 0.5 hPa), and (d, h) precipitation (contour interval: 1 mm d$^{-1}$, with additional contours for -0.5 and +0.5 mm d$^{-1}$ being inserted) for DJF(0/1) (left panels) and MAM(1) (right panels). Results are based on NCEP reanalysis data for six selected warm ENSO events and six cold events. Solid and dashed contours indicate positive and negative values, respectively.